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A complete Late Weichselian and Holocene record of aeolian coversands, drift sands and soils forced by climate change and human impact, Ossendrecht, the Netherlands

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Abstract

A stacked aeolian sequence with intercalated soils is presented from the southern Netherlands, which fully covers the Late Weichselian and Holocene periods. An integrated sedimentological (sedimentary structures, grain size), palynological (pollen) and dating approach (radiocarbon, optically stimulated luminescence (OSL)) was applied to unravel climatic and human forcing factors. The dating results of soils and sediments are compatible, and no large hiatuses between the radiocarbon-dated top of the soils and OSL-dated overlying sands were observed. It is argued that the peaty top of wet-type podzols can be used for reliable radiocarbon dating. This study reveals more phases than previously known of landscape stability (Usselo Soil and two podzol soils) and instability (Younger Coversand I and II, two drift-sand units) that are related to Late Weichselian climate change and Holocene human occupation. Regional aeolian deposition in source-bordering (river) dunes (Younger Coversand II) took place in the second part of the Younger Dryas, after 12.3 ka cal. BP, implying a delayed response to Younger Dryas cooling, vegetation cover decline and river pattern change of the Scheldt. The onset of podzolisation and development of ericaceous vegetation occurred prior to the introduction of Neolithic farming, which is earlier than previously assumed. Early podzolisation was followed by a short phase of local drift-sand deposition, at c.5500 cal. BP, that possibly relates to agriculture. Strong human impact on the landscape by deforestation and agriculture resulted in a second phase of widespread drift-sand deposition covering the younger podzol soil after AD 1000.

Introduction

Phases of aeolian deposition and soil formation during the Late Weichselian have been associated with climate and climate change (Van der Hammen & Wijnstra, 1971). Cold conditions and sand deposition occurred during the Late Pleniglacial, Older Dryas and Younger Dryas periods, while warmer conditions during the Bølling, Allerød and Holocene resulted in soil formation. However, a strict correlation between climate change and aeolian activity has been questioned, as aeolian deposition seems to continue over the Late Pleniglacial to Late Glacial boundary (Kasse, 1997, 2002; Vandenberghe et al., 2013). The classic aeolian record of the Late Weichselian in the Netherlands (Older Coversand I, Beuningen Gravel Bed, Older Coversand II, Lower Loamy Bed, Younger Coversand I, Usselo Soil, Younger Coversand II) (Van der Hammen 1951, 1971; Vandenberghe et al., 2013 and references therein) has been recognised in large parts of northwest, central and southwestern Europe (e.g. Kolstrup, 1980; Kolstrup et al., 1990; Kozarski & Nowaczyk, 1991; Bateman, 1995; Janotta et al., 1997; Koster, 2005; Kasse et al., 2007, 2018; Derese, 2011; Costas et al., 2012; Sitzia et al., 2015, 2017).

Widespread aeolian activity probably stopped or sharply decreased during the onset of the Holocene because of the rapid expansion of a protective vegetation cover (Hoek, 1997, 2001; Vandenberghe et al., 2013; Kasse et al., 2018). In the Early Holocene, Mesolithic hunter-gatherer activity, intentional forest fires and habitation resulted in early heathland development, soil acidification, podzolisation and local drift-sand formation (Van Geel et al., 2017; Sevink et al., 2018).

Middle Holocene drift-sand phases (c.9–3 ka cal. BP) were reported, although not frequently, from the Netherlands and Germany (Van Mourik et al., 1995, 2010, 2012; Tolksdorf & Kaiser, 2012; Sevink et al., 2013, 2018; Koster, 2017). The cause for these earlier drift-sand phases, whether climate or man, is still debated (Clemmensen et al., 2001; Tolksdorf & Kaiser, 2012; Sevink et al., 2013, 2018; Koster, 2017; Pierik et al., 2018). The Holocene is regarded as a climatically fairly stable period; however, climatic oscillations (e.g. 8.2 ka, 2.8 ka) are reported (Van Geel et al., 1996; Davis et al., 2003; Blaauw et al., 2004; Wiersma & Renssen, 2006;

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Hoek & Bos, 2007). Whether these oscillations and related changes in vegetation were strong enough to cause landscape degradation to initiate drift-sand formation is questionable (Pierik et al., 2018). On the other hand, the impact of man on the landscape during the Holocene increased, especially from the Neolithic period onwards when the start of agriculture resulted in local deforestation and expansion of heathlands (Janssen, 1972; Van Geel et al., 2017). In the loess region of Limburg the start of the Neolithic occurred at c.7300 cal. BP, while in the coversand regions of the southern Netherlands it occurred later (Verhart & Arts, 2005: 244–5). Neolithic occupation in the Maas valley in the southeastern Netherlands was reported from c.6400 cal. BP (4365 BC; Bos & Zuidhoff, 2015). During the Bronze Age, Iron Age and Roman period, population density increased, deforestation and agriculture further increased and the expansion of heather fields is reported (Groenewoudt et al., 2007; Bos & Zuidhoff, 2015; Van Beek et al., 2015). Following the Roman period, population decreased and reforestation occurred, especially in the southern Netherlands (Roymans & Gerritsen, 2002).

From the Middle Ages onwards, after AD 950–1150, aeolian activity resumed and culminated in extensive aeolian drift-sand areas at the end of the 18th century (Koster, 1978, 2009, 2010, 2017; Castel et al., 1989; Koster et al., 1993; Van Mourik et al., 1995, 2010; Janotta et al., 1997; Derese et al., 2010a; Jungerius & Riksen, 2010; Pierik et al., 2018). Although a climatic cause for drift-sand formation, due to drier conditions during the ‘climatic optimum’ of the Middle Ages or because of increased storminess during the Little Ice Age (Cunningham et al., 2011), has been suggested (e.g. Heidinga, 1984, 2010; Jungerius & Riksen, 2010), this drift-sand phase is generally ascribed to an increase of population density and pressure on the landscape (De Keyzer 2016; De Keyzer and Bateman 2018; Pierik et al., 2018). Deforestation and expansion of agriculture, grazing and burning practices, formation of roads and cattle/sheep drifts, and the use of plaggen as fertiliser led to the remobilisation of the Late Weichselian aeolian sands (Koster, 2009, 2010; Pierik et al., 2018). This classic view on drift-sand formation was recently modified by De Keyzer (2016) and De Keyzer and Bateman (2018), who stress the importance of socio-economic causes and the distribution of power of stakeholders in the use of the commons and measures to cope with drift-sand development.

The time control of coversand and drift-sand units was previously based on pollen analysis (biostratigraphy) and radiocarbon dates of intercalated Late Glacial and Holocene soils and peat (Van der Hammen 1951, 1971; Van der Hammen et al., 1967; Zagwijn, 1974; Castel et al., 1989; Van Geel et al., 1989; Van Mourik et al., 1995; Hoek, 1997). In this way, aeolian sand deposits were dated indirectly by dating of soils, assuming a minimal hiatus between the end of organic accumulation in the soil and the onset of sand deposition. However, radiocarbon dating of soils, to date the end or onset of underlying or overlying aeolian sand units, has given unsatisfactory results (Van Mourik et al., 1995, 2010, 2011; Koster, 2009; Wallinga et al., 2013; Pierik et al., 2018). Soil organic material is the cumulative product of the soil formation period, and the obtained ages are different in relation to the organic fraction used (Van Mourik et al., 1995). Therefore, radiocarbon dating of a buried soil does not necessarily represent the start of sand deposition and can yield too old ages for the overlying drift-sand activity. In the type locality of the Lutterzand area, the top of the Holocene podzol was radiocarbon-dated at 1550 ± 43 BP while deposition of the overlying drift sands started at 350 ± 30 years ago based on optically stimulated luminescence

(OSL) dating, implying a hiatus of c.1200 years (Vandenberghé et al., 2013). Furthermore, shorter and longer hiatuses may be present at the base, the top or within the sand units, since it is difficult to estimate the duration of sedimentary boundaries or erosional contacts. It has been stated that the so-called Younger Coversand II coincides with the Younger Dryas period (Van der Hammen, 1951; Vandenberghé, 1991; Isarin et al., 1997) as it overlies the Usselo Soil of Allerød age (Hoek, 1997: 43) and is covered by the Holocene podzol soil. However, time lags or delayed responses to climate change may be present at the base and top of Younger Dryas aeolian deposits, related to a gradual vegetational cover decline at the Allerød – Younger Dryas transition or vegetation increase at the Younger Dryas to Holocene boundary.

Direct dating of the sand units by OSL circumvents the problem of indirect dating based on radiocarbon dates of soils. Luminescence dating enables direct dating of aeolian sediments and phases of landscape instability. OSL dating of quartz has provided reliable ages for Late Weichselian and Holocene aeolian sediments in the northwest European lowlands (Stokes, 1991; Bateman, 1995, 1998; Janotta et al., 1997; Radtke & Janotta, 1998; Bateman and Van Huissteden 1999; Bateman et al., 2000; Vandenberghé et al., 2004, 2009, 2013; Kasse et al., 2007, 2018; Wallinga et al., 2007; Buylaert et al., 2009; Derese et al., 2009, 2010a, 2010b, 2012; Van Mourik et al., 2010; Bogemans & Vandenberghé, 2011; Sevink et al., 2013, 2018; Beerten et al., 2014).

Although aeolian deposits have been investigated frequently in the Netherlands, they often represent only a few phases. Stacked aeolian and soil sequences covering most of the Late Weichselian and Holocene periods, in which both luminescence and radiocarbon dating techniques have been applied (e.g. by Vandenberghé et al., 2004, 2013; Van Mourik et al., 2010; Wallinga et al., 2013), are still rare in the Netherlands and in western Noord Brabant. Multi-proxy studies of aeolian sand deposits in the Netherlands, in which lithological and palynological dating are integrated (e.g. by Sevink et al., 2013, 2018), are even more scarce up to date.

Exposure Boudewijn near Ossendrecht in the southern Netherlands represents a long record of Late Weichselian and Holocene aeolian activity and soil formation. Therefore this paper aims: (i) to reconstruct the sedimentary environments and their changes over time based on the geometry, sedimentary structures and grain size of the aeolian units; (ii) to explore the suitability of radiocarbon dating of soils by combination with the previously established OSL dating results (Fink, 2000; Vandenberghé et al., 2004) and to establish a reliable chronological framework of phases of aeolian activity and landscape instability; (iii) to reconstruct vegetation changes during the landscape stability phases by pollen analysis of intercalated soils; and (iv) to explore the role of climate and man in vegetation and landscape development and possible leads or lags to external forcings.

Geological and geomorphological setting

The investigated Boudewijn sand pit is located in the southwestern Netherlands, province Noord Brabant, east of the village of Ossendrecht (Fig. 1A). This region is situated on the uplifting Campine block that is separated from the Roer Valley Graben by the Feldbiss Fault zone and its northwestern continuation into the Gilze–Rijen fault zone (Van Balen et al., 2000, 2005).

Western Noord Brabant is characterised by two regions with strongly contrasting geomorphology separated by a sharp transition (Fig. 1B). The eastern part is c.15–25 m above sea level and

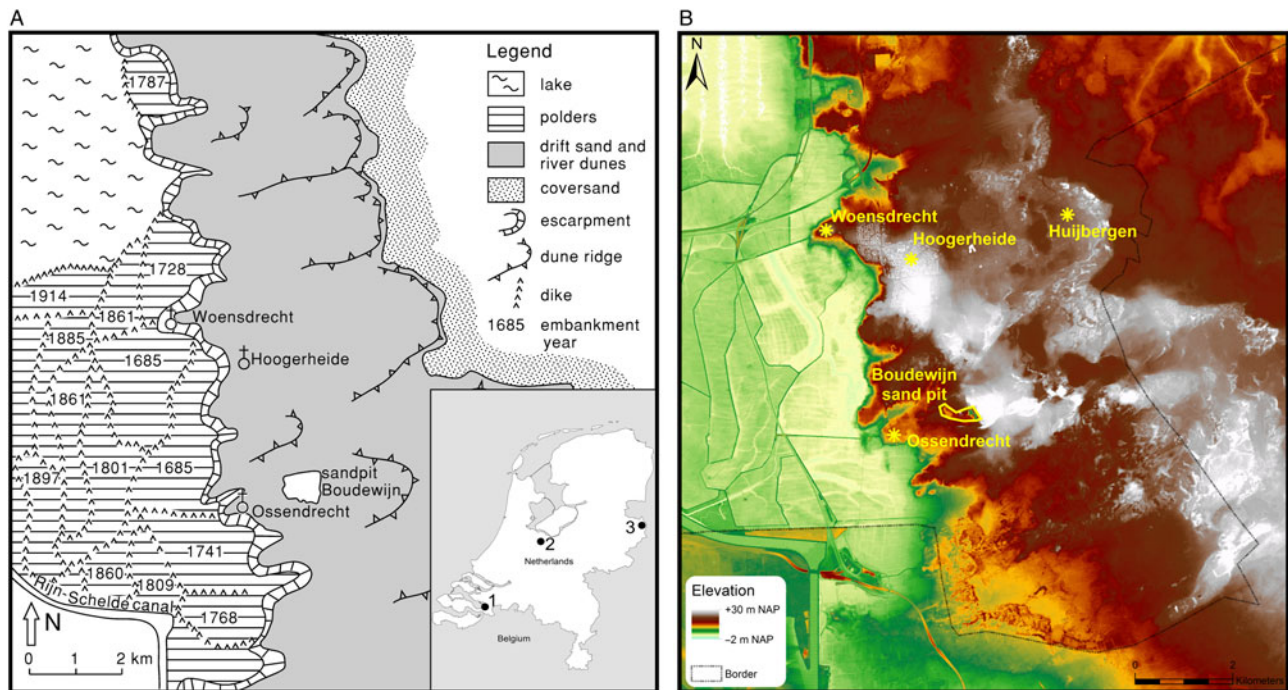


Fig. 1. (A) Location map and geomorphological setting of the investigated exposure Boudewijn near Ossendrecht in western Noord Brabant. Inset map of the Netherlands: 1. Ossendrecht, 2. Laarder Wasmeren (Sevink et al., 2013, 2018), 3. Lutterzand (Vandenberghe et al., 2013). (B) Digital elevation model of the study area showing the aeolian parabolic dune morphology east of Boudewijn and the escarpment towards the polder area in the west.

characterised by Late Pleistocene and Holocene dunes covered with pine forest (Brabantse Wal). The western part is around sea level and is characterised by polders with Late Holocene tidal flat and marsh deposits that have been reclaimed during the last centuries. The transition between the two landscapes is an almost 30 km long north–south-oriented escarpment between Antwerp in Belgium and Bergen op Zoom in the Netherlands. The scarp has been formed by Middle and Late Pleistocene fluvial erosion by the Scheldt river. In addition, concentrated ground-water seepage and associated transport of sand at the foot of the escarpment promoted erosion of the escarpment (Kasse, 1988; Westerhoff & Dobma, 1995). Faults, as a possible explanation of the scarp, have not been reported from the area, since subsurface Neogene units (e.g. the Oligocene Boom Clay) do not show an offset below the scarp (Zagwijn and Van Staaldin 1975, geological overzichtsprefiel CC').

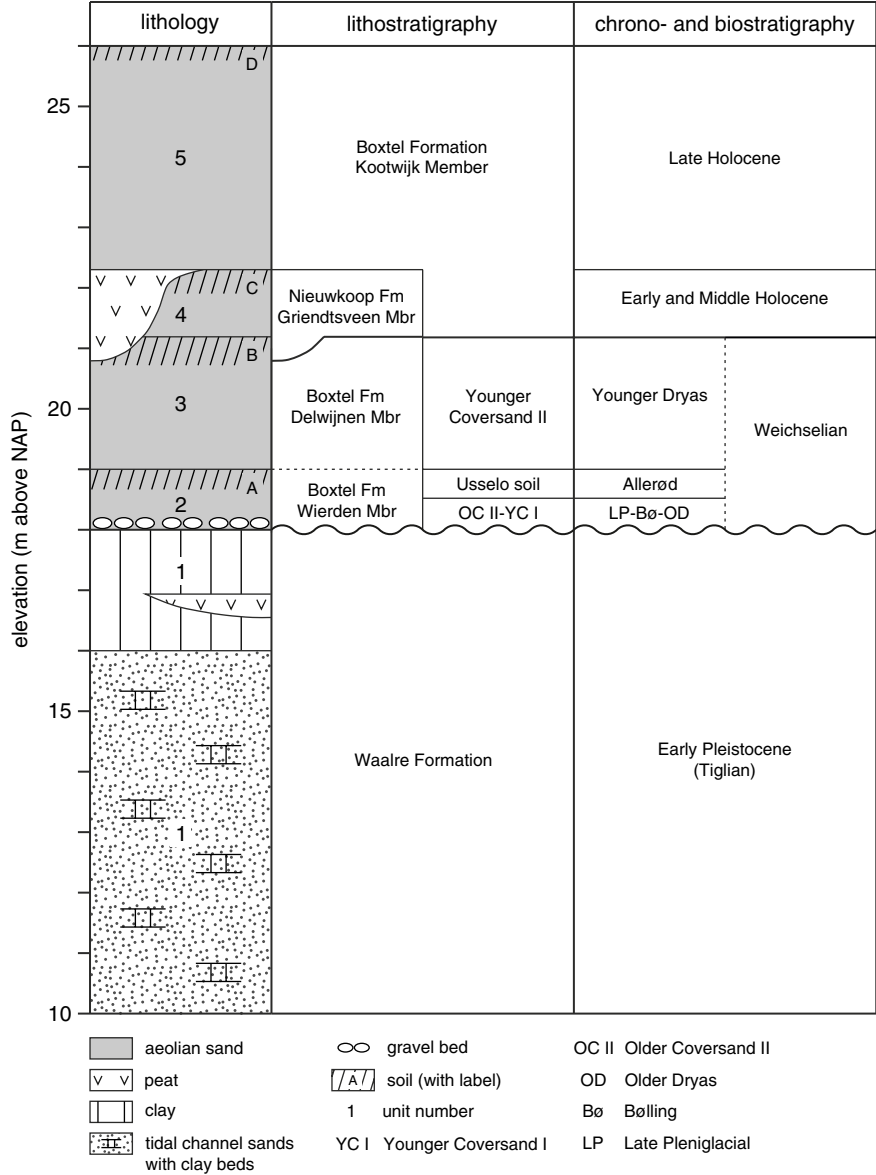
The Quaternary geology of western Noord Brabant and northern Belgium is characterised by Early Pleistocene tidal estuarine sediments (Fig. 2: unit 1) that are overlain by Late Pleistocene and Holocene mostly aeolian deposits (Fig. 2: units 2–5). The Early Pleistocene deposits dip in a northerly direction towards the subsiding North Sea basin (Kasse, 1988, 1990, 1993). They belong to the Waalre Formation of late Tiglian age (c.1.7–2.0 million years) (Westerhoff et al., 2008). The sediments were deposited at the southern margin of the North Sea basin in the confluence area of the Rhine and Scheldt rivers. The Waalre Formation (c.50 m thickness) consists of fine to medium white sands with thin discontinuous clay layers (flaser and wavy bedding), overlain by a 1–5 m thick clay bed (Fig. 2). The formation was previously interpreted as fluvial braided deposits (Van Dorsser, 1956; Damoiseaux, 1982). However, the sedimentary bedding points to tidal depositional environments (Kasse 1986) in accordance with conclusions by Dricot (1961) for the equivalent Campine/Weelde Formation in Belgium. An estuarine environment with laterally migrating tidal

channels in a warm-temperate climate during a sea-level high stand has been proposed (Kasse, 1988; Kasse & Bohncke, 2001). The bluish-grey clay layer at the top of the formation was formed in a tidal marsh, changing upward into a freshwater swamp with local alder peat formation (Fig. 2).

The Early Pleistocene Waalre Formation is separated from the overlying Late Pleistocene Boxtel Formation by an unconformity characterised by a thin gravel bed (Fig. 2). The gravel (2 mm to 5 cm) is partly broken by frost shattering and consists predominantly of well-rounded quartz and flint (Scheldt gravel association) (Zandstra, 1969). The gravel, derived from Tertiary marine deposits south of the study area in Belgium, was deposited by the northward-flowing, ancestral Scheldt system in the Stramproy Formation during the Early Pleistocene (Kasse, 1988; Westerhoff et al., 2008). Middle and Late Pleistocene uplift of the study area resulted in a northern shift of the North Sea basin hinge line. Subsequent erosion under periglacial climate conditions has eroded most of the Stramproy Formation, and the gravel was concentrated in an erosional residue on top of the Waalre Formation. In addition, the opening of the Dover Strait and the rearrangement of the drainage systems in the southern North Sea basin may have contributed to the Scheldt escarpment formation and deep erosion of the Flemish Valley west of the study area (Kasse, 1988; Gibbard, 1995; Cohen et al., 2014).

Previous research and age control of the site

Sand pit Boudewijn was investigated for more than 20 years (1984–2004) (Fig. 3A, B, C). In 2005 the exploitation of sand for the production of carbonate sandstone bricks stopped and pit faces degraded. The lithostratigraphy of the Early and Late Pleistocene sediments was studied by Kasse (1988). Later Schwan (1991) studied the sedimentology and sedimentary facies of the Weichselian



Late Glacial to Holocene aeolian succession. He presented an uncalibrated radiocarbon age for the Late Glacial Usselo Soil ($11,240 \pm 50$ BP) and two dates for the Holocene peat, which is time-equivalent with the Holocene podzol soil (9050 ± 45 BP for the base of the peat; 3000 ± 30 BP for the top of the peat) (Table 1). Fink (2000) and Vandenberghe et al. (2004) dated the Late Weichselian and Holocene aeolian sediments by OSL dating. Older OSL dates may be less reliable and less precise (larger standard deviation); however, they are incorporated in Figure 9 further below, in which the dating information is combined, as there are no systematic outliers related to a certain method. Georadar investigations of the aeolian deposits have been performed by Van Dam and Schlager (2000). Radiocarbon ages are presented in years BP, calibrated radiocarbon ages in cal. BP and OSL ages in ka.

Methods

Grain-size analyses were performed following the method described by Konert and Vandenberghe (1997). Samples were pre-treated with H_2O_2 and HCl to remove organic matter and

calcium carbonates. Grain-size distributions ranging from 0.1 to $2000 \mu m$ were measured with a Fritsch laser-diffraction instrument at the VU University Amsterdam.

Late Glacial Soil A and the two Holocene podzol soils B and C were sampled for pollen analysis and radiocarbon dating to reconstruct temporal vegetation changes that can be linked to climate change or human activity (Fig. 2). Local pollen assemblage zones are defined based on changes in species composition. They are correlated with the biochronostratigraphic zonation for the Late Glacial and Holocene in the Netherlands (Janssen, 1974; Van Geel et al., 1981; Hoek, 1997; Bos et al., 2006), which provides indirect dating for deposition of the aeolian units. The soils (A–E–B–C horizons) were sampled in the field in metal boxes ($40 \times 10 \times 10$ cm) and described in detail in the laboratory. Sample location was based on lithology and lithological transitions. The pollen samples were prepared following standard procedures (Faegri & Iversen, 1989). The pollen slides were examined using a Zeiss Axioskop 50 light microscope with a magnification of $630\times$ with and without phase contrast. Pollen was determined using the pollen key of Moore et al. (1991). The pollen sum includes tree taxa,

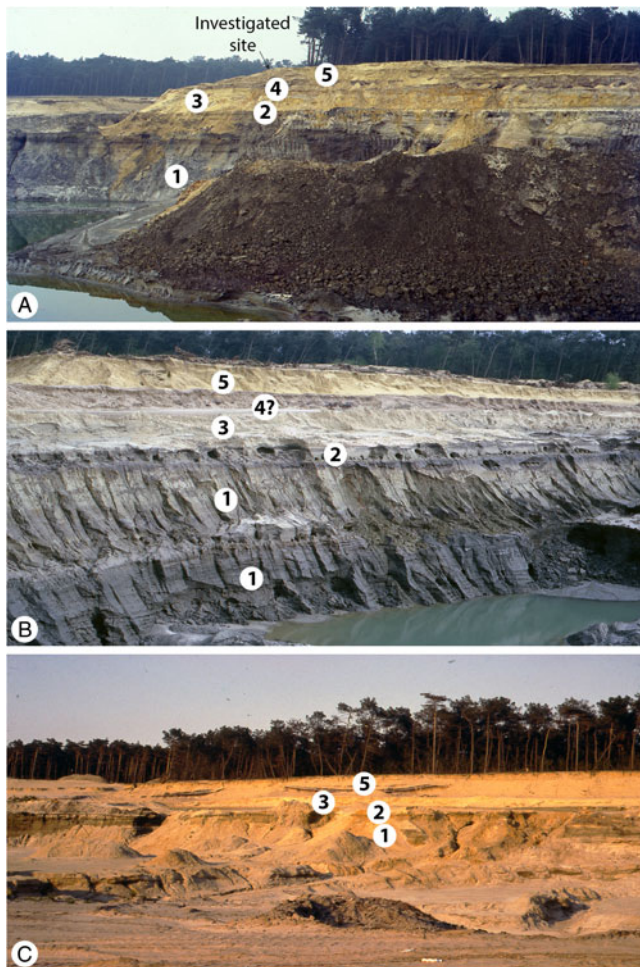


Fig. 3. Overview of the lithological succession in exposure Boudewijn during the exploitation period. (A) 1997 view to the north, section c.15 m high; (B) 1999 view to the northeast, section c.15 m high; (C) 1990 view to the east, section c.8 m high. Unit 1 = Waalre Formation; Unit 2 = Older Coversand II / Younger Coversand I and Usselo Soil; Unit 3 = Younger Coversand II; Unit 4 = older Drift Sand and podzol soils; 5 = Drift Sands.

shrubs and upland herbs. Wetland herbs, aquatics and Cyperaceae are excluded from the pollen sum. Pollen diagrams are constructed using TILIA software in which taxa are arranged in ecological groups (Grimm, 1992).

Despite the known problems related to radiocarbon dating of soils (Van Mourik et al., 1995; Pierik et al., 2018), samples were taken at selected sites to investigate the reliability of soil dates in comparison with OSL dates of the overlying sand units. The aim of radiocarbon dating is to date the final moment of soil formation in order to establish the (maximum) age of the onset of aeolian deposition. The dates are based on bulk and macro remains samples (Table 1). Because dating of soils is often problematic due to old carbon in the soils, sampling locations were carefully selected in the field in plastic bags. Wood remains and peat were sampled from the top of the Usselo Soil. Humic, loamy fine sand with many macro remains (twigs) was sampled from the base of the Younger Coversand II unit. The podzols show lateral variability with humic-iron podzols on the higher locations and peaty podzols in palaeo-depressions. The uppermost top of the peaty podzols was selected for dating as the organic material accumulated on the wet surface and was less affected by bioturbation and incorporation of older

carbon from below. The top of podzol B at site OSD99-4 consisted of well-preserved sandy sedge peat indicating the onset of aeolian deposition of unit 4. Radiocarbon dates mentioned in the text are calibrated using Oxcal 4.3 (Bronk Ramsey, 2009; Reimer et al., 2013).

Results

Description of the Late Weichselian and Holocene units

The upper 8 m of the exposure (units 2, 3, 4, 5) consist of grey and yellow, fine (63–250 µm) to medium (250–500 µm) aeolian sands with intercalated soils that belong to the Bostel Formation (Wierden Member, Delwijnen Member, Kootwijk Member) of late Weichselian and Holocene age (Fig. 2).

The areal distribution of the units and soils was established by yearly visits over the exploitation period. Units 2, 3 and 5 and Soil A (Usselo Soil) were always observed and extend over several hundred metres or more (Figs 2 and 3). Generally one podzol soil was found separating units 3 and 5, except where the podzol was deflated prior to the deposition of unit 5. Unit 4 and podzol soils B and C were observed locally over c.100 m. Outside the distribution area of unit 4, podzol soils B and C merge into one podzol.

Unit 2 (Older Coversand II / Younger Coversand I)

Unit 2 is an extensive, 0.2–1 m thick, unit that overlies the clay of the Waalre Formation. The lower boundary is sharp with some fine gravel, and locally periglacial structures (wedge cast) were observed (Fig. 4A).

Unit 2 consists of well-sorted loamy fine sands with mean grain size of 210 µm ($n = 1$) (Fig. 5). The upper part is finer-grained and poorly sorted, with a mean grain size of 127 µm and a high clay and silt fraction (23%). Wavy horizontal bedding, locally disturbed by cryogenic deformations, has been observed (Fig. 4B), but sedimentary structures are frequently obscured because of water-saturated conditions related to the impermeable clay below.

Soil A (Usselo Soil and basal part of Younger Coversand II)

Soil A is a multi-genetic soil complex. The lower part developed in the top of unit 2. It consists of a c.10 cm thick, dark-grey to black, humic Ah soil horizon (Usselo Soil), that locally grades laterally into a (involved) peaty bed (Fig. 4B). The upper part of soil complex A consists of slightly organic, loamy, laminated fine sands that reflect the onset of deposition of unit 3.

Samples for pollen analysis were taken from soil complex A (Fig. 6A, B) and the result is presented in Figure 7A. Three local pollen zones are distinguished. Zone 1 (34–36 cm) is characterised by high *Betula* values and some upland herbs. In zone 2 (29.5–34 cm) *Betula* has declined and *Pinus* and ferns are dominant. Zone 3 (0–29.5 cm: organic loamy fine sand) shows, compared to zone 2, a strong decline in *Pinus* and an increase of *Juniperus* and *Salix*. Ericaceae and dryland herbs are important, while high Cyperaceae and *Sphagnum* values indicate local wet conditions. *Empetrum* and *Armeria* are present in low values. *Alnus*, *Ilex* and *Tsuga* show an increase in this zone.

Radiocarbon dates of wood fragments from the black Ah horizon (Usselo Soil) provided uncalibrated ages of $10,870 \pm 40$, $10,900 \pm 30$ and $11,060 \pm 60$ BP (Table 1), which point to a late Allerød age of this soil (Vandenberghe et al., 2004). Radiocarbon dates from the overlying loamy, organic fine sand (base of Younger Coversand II) yield uncalibrated ages of $10,460 \pm 180$ (cf) and $10,680 \pm 100$ BP (ff) for

Table 1. Radiocarbon dates of the organic units and soils in the Late Weichselian and Holocene aeolian succession. Radiocarbon dates mentioned in the text were calibrated to calendar years before AD 2000 using Oxcal 4.3 (Bronk Ramsey, 2009; Reimer et al., 2013) cf is coarse fraction; ff is fine fraction

Field code	Elevation (m NAP ^a)	Material	GrN number ^b	¹⁴ C age BP	Calendar age (years before AD 2000)	Reference
OSD 99-3	22.2	Slightly sandy brown peat (top soil C)	25849	1000 ± 30 (cf)	1017–949 (71.0%) 918–872 (19.2%) 865–848 (5.2%)	This study
			25937	890 ± 30 (ff)	959–892 (35.9%) 884–782 (59.5%)	
OSD 99-4b	21.1	Sandy sedge peat (top soil B)	25850	4940 ± 50 (cf)	5913–5878 (4.3%) 5804–5639 (91.1%)	This study
			25749	4940 ± 60 (ff)	5939–5860 (10.6%) 5814–5637 (84.8%)	
OSD 99-5e	19.3	Slightly organic, loamy sand at the base of Younger Coversand II	25852	10450 ± 190 (cf)	12777–11750 (94.8%) 11722–11691 (0.6%)	Vandenbergh et al. (2004)
			25870	11080 ± 180 (ff)	13334–12745 (95.4%)	
OSD 99-5c	19.1	Organic, loamy sand at the base of Younger Coversand II	25851	10460 ± 180 (cf)	12775–11800 (95.2%) 11784–11773 (0.2%)	Vandenbergh et al. (2004)
			25948	10680 ± 100 (ff)	12799–12461 (95.4%)	
OSD 99-22	18.9	Wood from top of Usselo Soil	25871	10900 ± 30 (cf)	12862–12756 (95.4%)	Vandenbergh et al. (2004)
			25853	10870 ± 40 (ff)	12856–12746 (95.4%)	
			25940	11060 ± 60 (peat)	13121–12826 (95.4%)	
III	—	Top of peat layer	12745	3000 ± 30	3377–3348 (6.0%) 3304–3126 (89.4%)	Schwan (1991: 157)
II	—	Base of peat layer	12744	9050 ± 45	10326–10211 (95.4%)	Schwan (1991)
I	—	Usselo peat or soil	12743	11240 ± 50	13265–13070 (95.4%)	Schwan (1991)

^aNormal Amsterdam Level.

^bLabcode of radiocarbon dating facility of Groningen University.

the lower sample; $10,450 \pm 190$ (cf) and $11,080 \pm 180$ (ff) for the upper sample (Table 1; Fig. 7A). The fresh brown organic material formed *in situ* and was not reworked from the underlying darker Usselo soil. Therefore, it provides the age of the onset of aeolian sedimentation of unit 3 (Younger Coversand II).

Unit 3 (Younger Coversand II)

Unit 3 overlies soil complex A and is generally 2 m thick, but the maximum thickness is unknown due to later deflation of the higher parts of unit 3 (Figs 2 and 3). Unit 3 is present over the whole exposed area and shows an overall coarsening-upward trend from c.100 to 300 μm (Fig. 5A). The lower 25 cm consists of greyish-brown, humic, laminated, loamy fine sands (mean 122 μm , 19–25% clay and silt; $n=2$). Loam and organic matter content decrease upward which is related to increased aeolian sand deposition (Fig. 6B). The main part of unit 3 consists of light-grey to yellow medium sands with gleyic mottles and a mean grain size of 267 μm ($n=7$) (Fig. 5A). The clay and silt fraction is low, at 1–3%, and a prominent coarse sand fraction larger than 500 μm of 8–17% is present (Fig. 5B).

In the lower part of unit 3, wavy horizontal bedding, crinkly lamination and adhesion ripple bedding are the dominant bedding types (Figs 4C and 8A). The upper part of the sequence is dominated by plane horizontal bedding and low-angle cross bedding (Fig. 4D). High-angle, dune slipface cross-bedding is rarely observed (Fig. 4E).

Periglacial structures like frost cracks, ice-wedge casts or sand wedges have not been observed, but locally metre-scale large involutions were found (Fig. 4F).

Soil B (podzol)

Soil B has formed in the upper part of unit 3 (Younger Coversand II). A humic-iron podzol was found on higher well-drained dunes, and a peaty podzol was found in the former dune slacks (Fig. 6A, C, D). The podzol formed in and follows the undulating morphology of unit 3, which indicates that unit 3 was deposited in low dunes not exceeding 5 m in height (Fig. 3C).

Podzol soil B was sampled at site OSD99-4 (Fig. 6A, C). The results of the pollen analysis are presented in Figure 7B. Two pollen zones are distinguished. Zone 1 (23–25 cm) is dominated by high arboreal pollen values, mainly of *Corylus* and *Pinus*, and by Ericaceae (Boreal period). The start of zone 2 (7–23 cm) is characterised by the first presence of *Alnus*, *Quercus*, *Tilia*, *Ulmus* (Atlantic period). Somewhat higher in this zone, *Fraxinus* appears and increases in value. Ericaceae are present throughout the zone, and Cyperaceae and Poaceae increase upward. Some dryland herbs like *Artemisia*, Chenopodiaceae, Asteraceae and Brassicaceae are present that possibly indicate agricultural activity.

The sandy sedge peat at the top of podzol soil B was radiocarbon-dated at 4940 ± 50 BP (cf) and 4940 ± 60 BP (ff) (Table 1).

Unit 4 (older Drift Sand)

Unit 4, intercalated between podzol soils B and C, has a thickness of maximal 1 m (Fig. 6A, D). This unit was observed locally over a distance of c.100 m. Unit 4 consists of light-grey to yellow, medium sand with a mean grain size of 303 μm ($n=3$) and a weak coarsening-upward trend (Fig. 5A). Grain-size distributions resemble unit 3. The fine fraction smaller than 63 μm is almost absent (less than 1–2%), but a distinct coarse sand fraction larger than 500 μm is present (8–21%) (Fig. 5B).

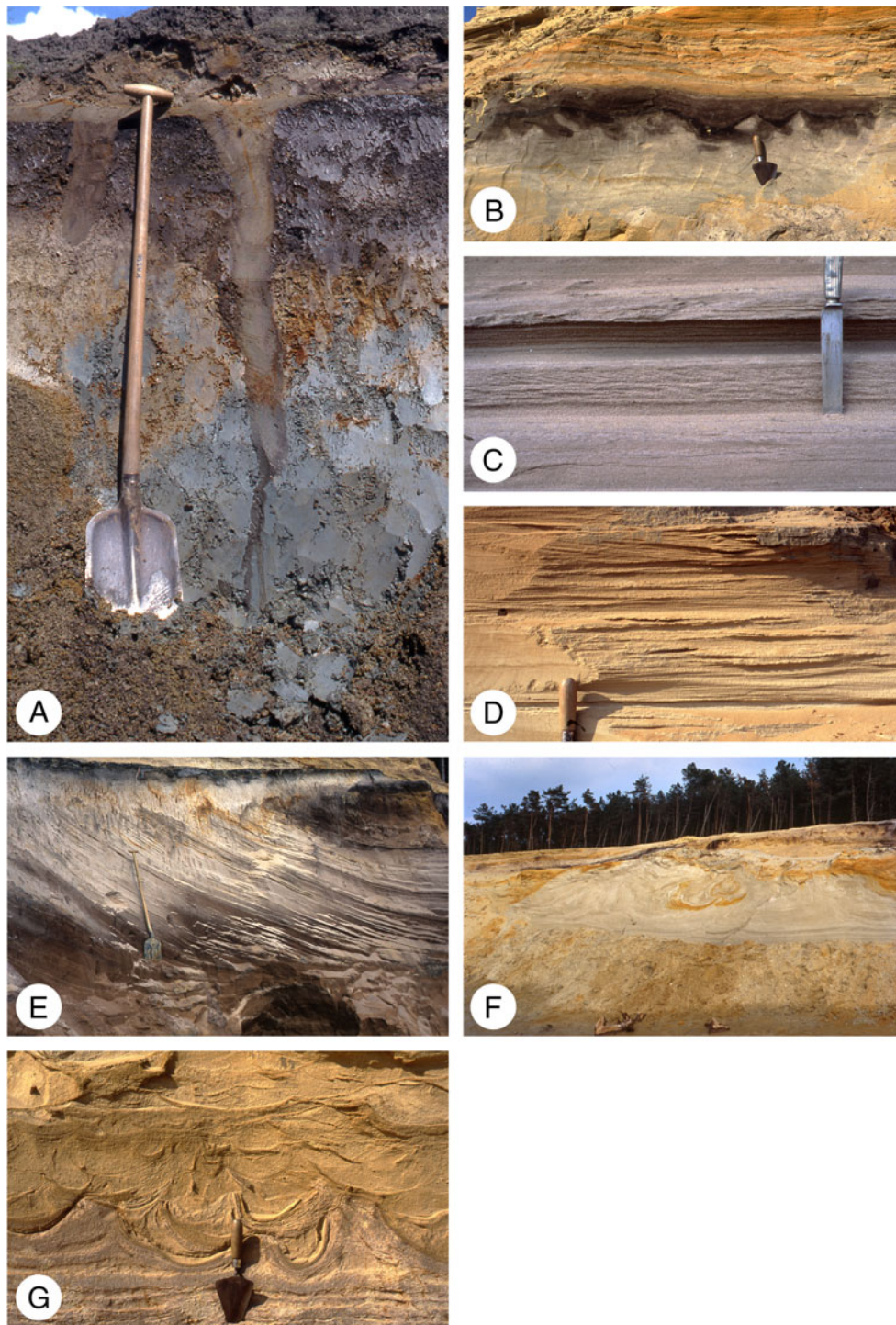


Fig. 4. Photographs of the units. (A) Large frost crack in the top of the Waalre Formation. (B) Cryoturbated peaty Usselo Soil. (C) Alternating horizontal and adhesion ripple cross-lamination in the wet-aeolian lower part of unit 3 (Younger Coversand II). (D) Horizontal and low-angle cross-bedding in the dry-aeolian upper part of unit 3 (Younger Coversand II). (E) Large-scale dune-slipface cross-bedding in unit 3 (Younger Coversand II) in which the Holocene podzol has developed with thick eluviation horizon following the sedimentary bedding. (F) Local large-scale involutions in unit 3 (Younger Coversand II); note black Holocene podzol soil in the upper part of the section. (G) Several generations of small-scale involutions at the transition from wet to dry aeolian deposition in unit 5 (Drift Sand).

Bedding types are dominated by horizontal bedding and low-angle cross-bedding (Fig. 6A, D).

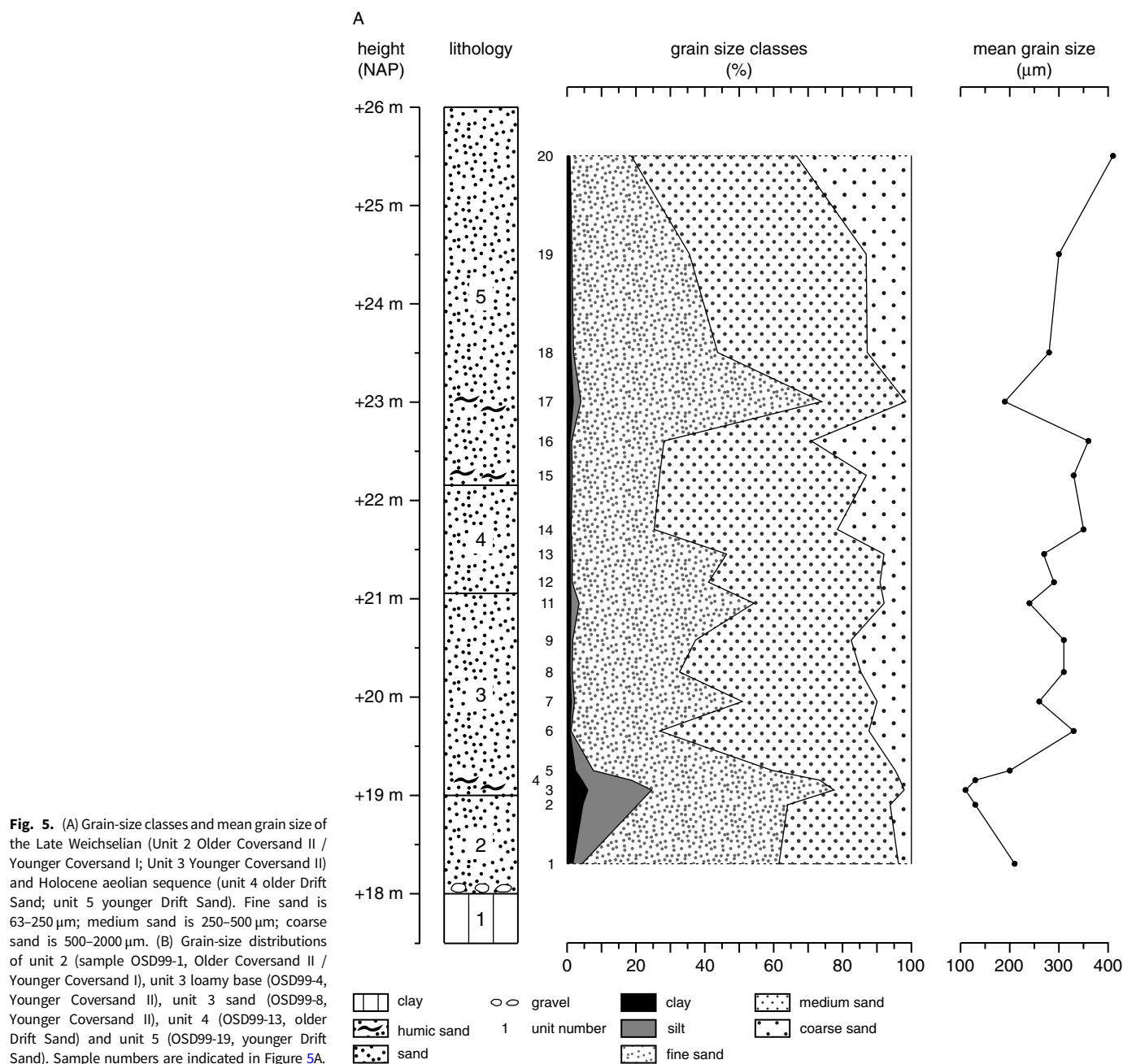
Soil C (podzol)

Soil C developed in the upper part of unit 4. Like soil B, the podzol shows a lateral change from a well-drained humic-iron podzol to a lower-lying peaty podzol (Fig. 6A). Site OSD99-3 has been sampled for pollen analysis and radiocarbon dating (Fig. 6C).

The pollen diagram of podzol soil C is presented in Figure 7C. Two pollen zones are distinguished separated by an interval poor

in pollen. Zone 1 (20–35 cm) is characterised by *Alnus*, *Corylus*, *Quercus*, *Tilia* and *Ulmus* (Atlantic–Subboreal period). *Fagus* and *Carpinus* have not been observed. Ericaceae values are high, and dryland herbs like Chenopodiaceae, Asteraceae, Brassicaceae and *Jasione* are present but Cerealia have not been found.

Zone 2 (4–6 cm) is characterised by low arboreal pollen values and high Ericaceae values. *Alnus*, *Betula*, *Corylus* and *Quercus* strongly decline compared to zone 1, while *Ulmus* and *Tilia* are scarce or even absent. *Fagus* and *Carpinus* are present for the first time. Upland herbs are well presented with Chenopodiaceae,



Asteraceae, Brassicaceae and *Rumex*. *Cerealia*, *Centaurea*-type and *Plantago* are characteristic elements of this zone (Subatlantic period).

Uncalibrated radiocarbon dates of 1000 ± 30 BP (cf) and 890 ± 30 BP (ff) from the peaty top of podzol soil C (Table 1; Fig. 7C: interval 3–5 cm) provide a maximum age for the deposition of drift-sand unit 5.

Unit 5 (younger Drift Sand) and Soil D (micropodzol)

Unit 5 overlies podzol soil C and was observed in all pit faces over the years of exploitation (Figs 3 and 6A). Its thickness ranges from 1 to 5 m. A weakly developed soil or micropodzol (Dutch: *Duinvaaggrond*) occurs at the present-day surface.

Unit 5 consists of brownish-grey to yellow, medium sand with a mean grain size of $312 \mu\text{m}$ ($n = 6$) (Fig. 5A). The lower brownish-

grey part fills in former depressions of the landscape (Fig. 6C). A coarsening-upward trend from 200 to $400 \mu\text{m}$ is visible in the upper part of the unit (Fig. 5A). The sediments have generally less than 2% clay and silt fraction ($<63 \mu\text{m}$). The coarse fraction larger than $500 \mu\text{m}$ varies between 13% and 33% and reaches the highest values of the complete aeolian succession (Fig. 5B). One sample (17) forms an exception to the general grain-size characteristics. It was taken at the transition from greyish to yellow sand and contained many crinkly humic laminae. It is finer-grained (mean $186 \mu\text{m}$), with higher fine fraction (4%) and lower coarse fraction (2%).

Wavy horizontal bedding, crinkly lamination and adhesion ripple bedding are the dominant bedding types in the lower brownish-grey part of unit 5 (Fig. 8B). The upper part of the sequence is dominated by plane horizontal bedding and low-angle cross-bedding (Fig. 6A and C). High-angle cross-bedding, indicating dune slipface deposition, is rarely observed.

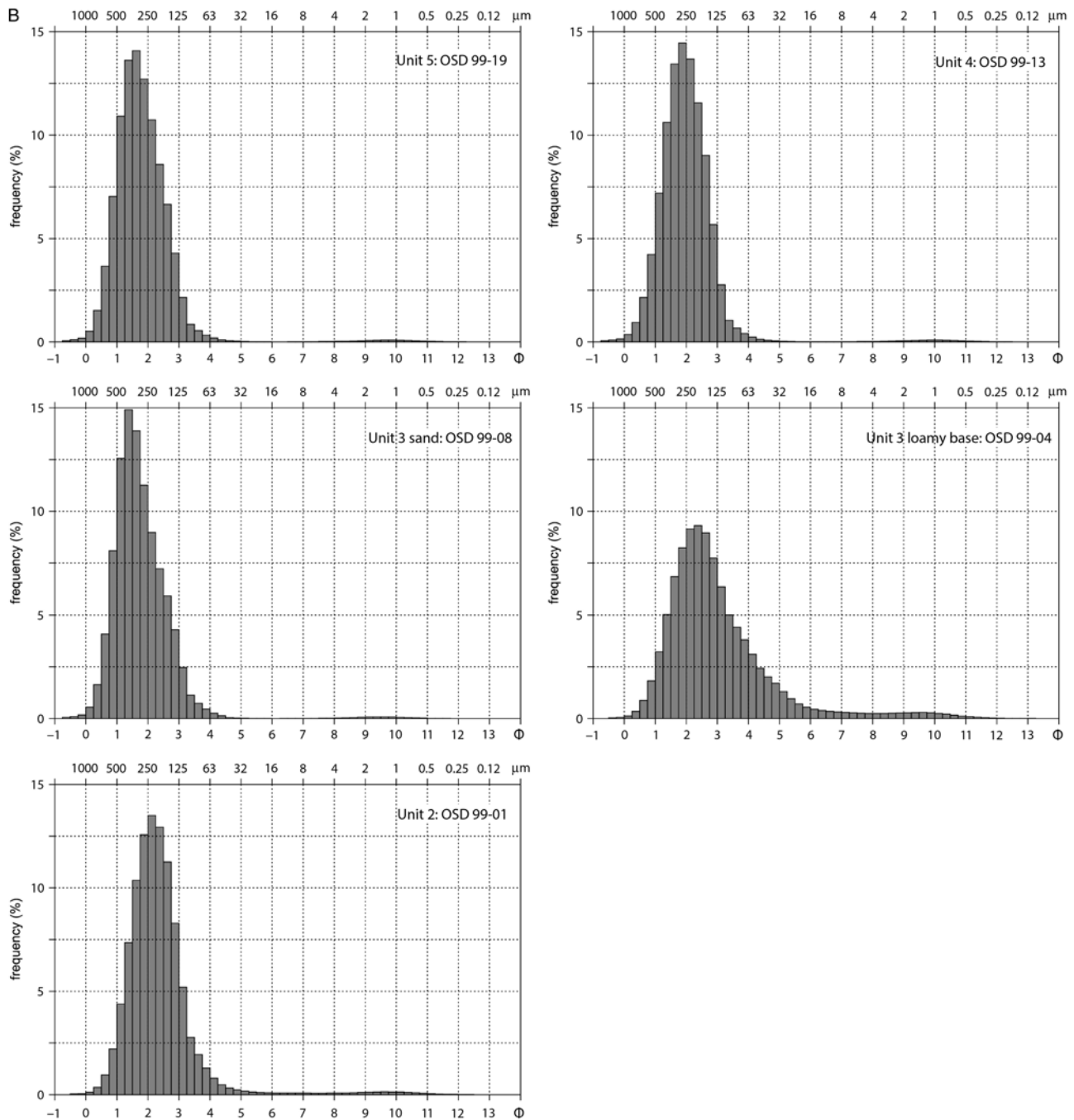


Fig. 5. Continued.

Decimetre-scale involutions are frequently found in unit 5, especially in the brownish-grey lower part and at the transition to the yellow upper part (Figs 4G and 8B). Several phases of loading followed by renewed sedimentation are present.

Interpretation and discussion

During 20 years of exploitation, exposure Boudewijn revealed the presence of four fine to medium-grained aeolian sand units with intercalated soils of late Weichselian and Holocene age (Figs 2 and 3). The results are summarised in Figure 9 in which the new results are combined with previous age determinations (Schwan,

1991; Fink, 2000; Vandenberghe et al., 2004). The local pollen zones of the soils are correlated with the well-established biostratigraphy of the Late Glacial (Hoek, 1997, 2001) and the Holocene (Janssen, 1974; Van Geel et al., 1981; Bos & Zuidhoff, 2015).

General sedimentary evolution

The grain-size analysis shows an overall coarsening-upward trend, with fine sands (mean 169 μm) in unit 2 grading upward into medium sands (200–400 μm) in unit 5 (Figs 5 and 9). Superimposed on this general trend, smaller-scale coarsening-upward sequences are visible within units 3, 4 and 5. These coarsening-upward trends on different scales can be explained by

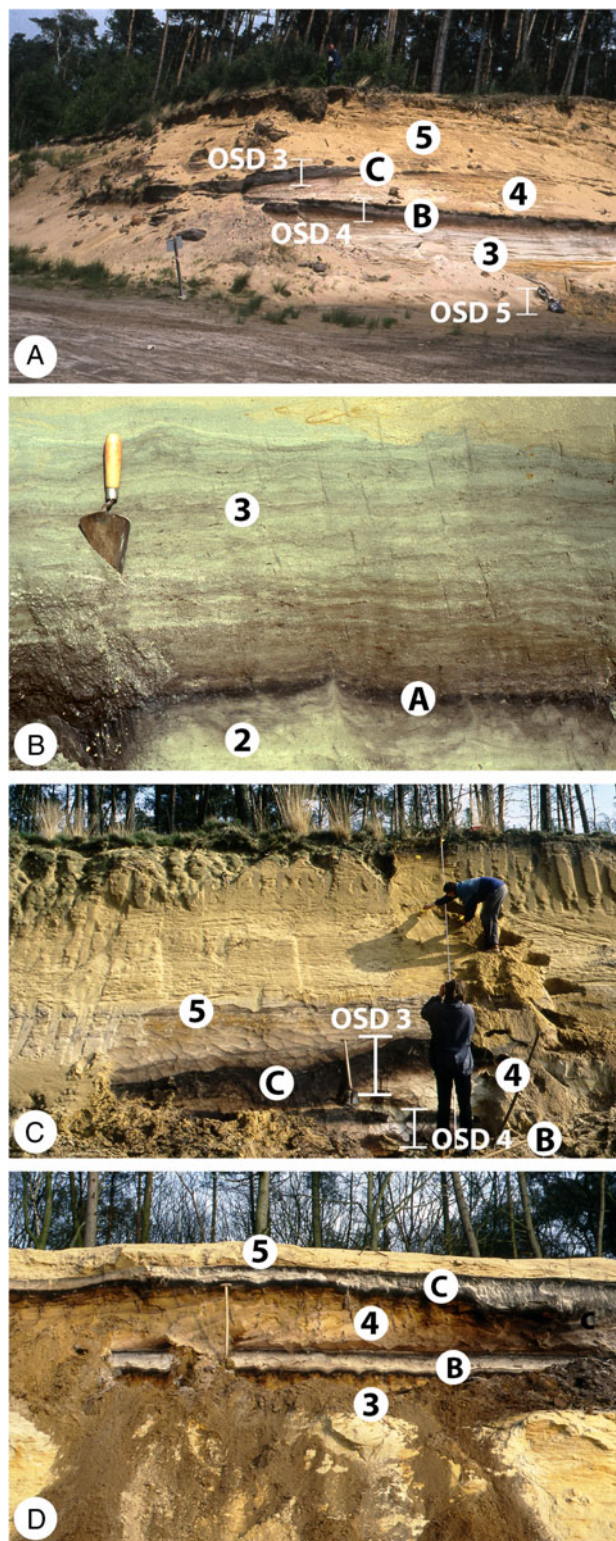


Fig. 6. (A) Location of the investigated pollen profiles OSD3 (soil C), OSD4 (soil B) and OSD5 (soil A); lithological units are indicated by numbers; for location see Figure 3A. (B) Sedimentary succession at pollen profile OSD5 with the Usselo Soil (darkest colour) at the base and laminated organic fine sand in the lower part of unit 3 (Younger Coversand II) near the trowel; note the gradual decrease in organic matter and increase in sand content reflecting the onset of aeolian deposition of unit 3. (C) Location of pollen profiles OSD3 (soil C) and OSD4 (podzol B); note that the two soils merge to the left into one soil and the upper soil changes laterally from a podzol to a peaty soil. (D) Two well-developed humic-iron podzols and intercalated older Drift Sand unit 4 (eastern pitface, c.50 m east of OSD3/4/5).

changes in the aeolian depositional environment (see discussion below). In the lower finer-grained parts of the aeolian units 2, 3 and 5, the presence of adhesion ripple bedding and wavy to crinkly lamination points to wet-aeolian deposition (Hunter, 1980; Ahlbrandt & Fryberger, 1982; Ruegg, 1983: facies B; Schwan, 1986, 1988: facies 3; Kasse, 2002: facies 2) (Figs 4B, C and 8). The sand was probably deposited from the traction load and the silty fine sand by adhesion of the suspension load on a moist to wet depositional surface. According to Schwan (1986), the sand and silty fine sand might reflect alternating deposition attributed to (possibly seasonal) variations in wind velocity. Storm events resulted in sand deposition while during fair weather periods silt settled from suspension and adhered on a damp surface. The well-developed palaeo-wetness indicators in the Boudewijn sand pit, despite its high position in the landscape, are related to the underlying impermeable clay of the Waalre Formation leading to a perched water table.

The upper parts of units 3, 4 and 5 are dominated by horizontal lamination and low-angle cross bedding, that indicate aeolian deposition in plane beds or by wind ripples on a dry surface of a sand sheet or in low dunes (Hunter, 1977; Fryberger et al., 1979; Ruegg, 1983: facies A; Schwan, 1986, 1988: facies 2; Kasse, 2002: facies 3) (Figs 4D and 6C).

The changes in grain size and sedimentary bedding in the units (and in the succession as a whole) indicate that changes from wet to dry aeolian depositional environments (units 3 and 5) occurred. Such vertical facies changes and drying-up sequences have previously been related to environmental changes in climate, permafrost or wind regime (Schwan, 1986; Koster, 1988: table 4; Stapert & Veenstra, 1988: 11–12). However, evidence for permafrost (ice-wedge casts) was not encountered in the aeolian sequence of exposure Boudewijn, nor in equivalent units in the Netherlands (Maarleveld, 1976; Kasse, 2002). These drying-up sequences can be explained by system-intrinsic changes in topography and drainage conditions by the accumulation of wind-blown sand and dune migration over wet produne or interdune areas caused by the underlying impermeable clay (Ahlbrandt & Fryberger, 1982; Schwan 1988: 224; 1991). The increasing thickness and relief during the deposition of the sediment bodies (units) resulted in a decrease of the soil moisture content and a drying-up sequence. This holds true for the individual units but also for the stacked up to 8 m thick aeolian complex overlying the Brabantse Wal (Fig. 2).

Late Pleniglacial and early Late Glacial coversand deposition (Unit 2: Older Coversand II / Younger Coversand I)

Some periglacial deformation structures were found in the top of the clay of the Waalre Formation (unit 1), separated from unit 2 by an erosional unconformity (Fig. 4A). The scarcity of periglacial deformation structures in the top of the clay, while they are reported more regularly to the east (Kasse, 1988, e.g. exposures Wouwse Plantage, Meerle), may be related to dry subsoil conditions below the clay layer, because of the low groundwater table in this region close to the Brabantse Wal escarpment. Apparently, ice-rich permafrost did not develop in the clay and sand of the Waalre Formation during the Late Pleniglacial, and periglacial loading due to melting of the permafrost did not occur.

Unit 2 is generally free from periglacial deformations. Ice-wedge casts and large-scale loading structures are generally interpreted as the result of permafrost associated with the Last Glacial Maximum (LGM) (Vandenberghe, 1985; Vandenberghe et al., 2013). Previous OSL dating of sand unit 2 provided direct dates ranging between 14.6 ± 0.9 ka and 15.3 ± 0.9 ka (Vandenberghe

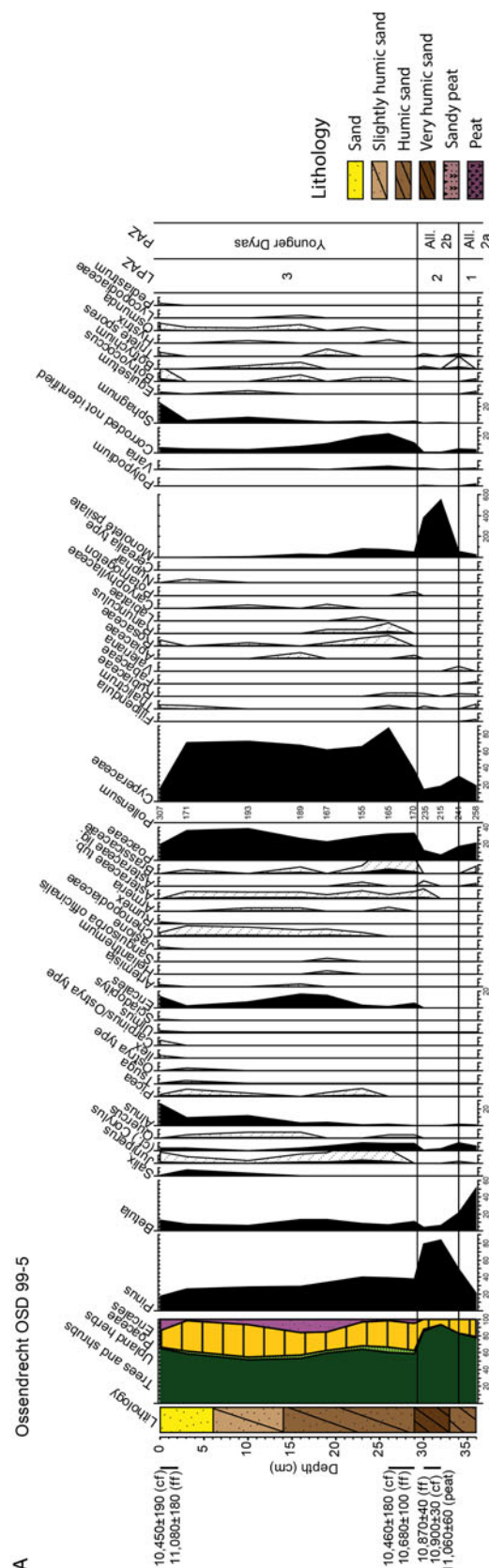


Fig. 7. (A) Pollen diagram OSD99-5 from soil complex A (Usselo Soil and fine-grained basal part of Younger Coversand II). Note the sharp drop in *Pinus* and increase of upland herbs and Poaceae at the Allerød – Younger Dryas transition. (B) Pollen diagram OSD99-4 from lower podzol B showing the Early to Middle Holocene vegetation development, early presence of heathland (*Ericaceae*) and weak indications of human activity (Poaceae, upland herbs). (C) Pollen diagram OSD99-3 from upper podzol C showing the Middle to Late Holocene vegetation development. Note the high presence of *Ericaceae* and *Cerealia* in the upper part, indicating strong human impact on the vegetation.





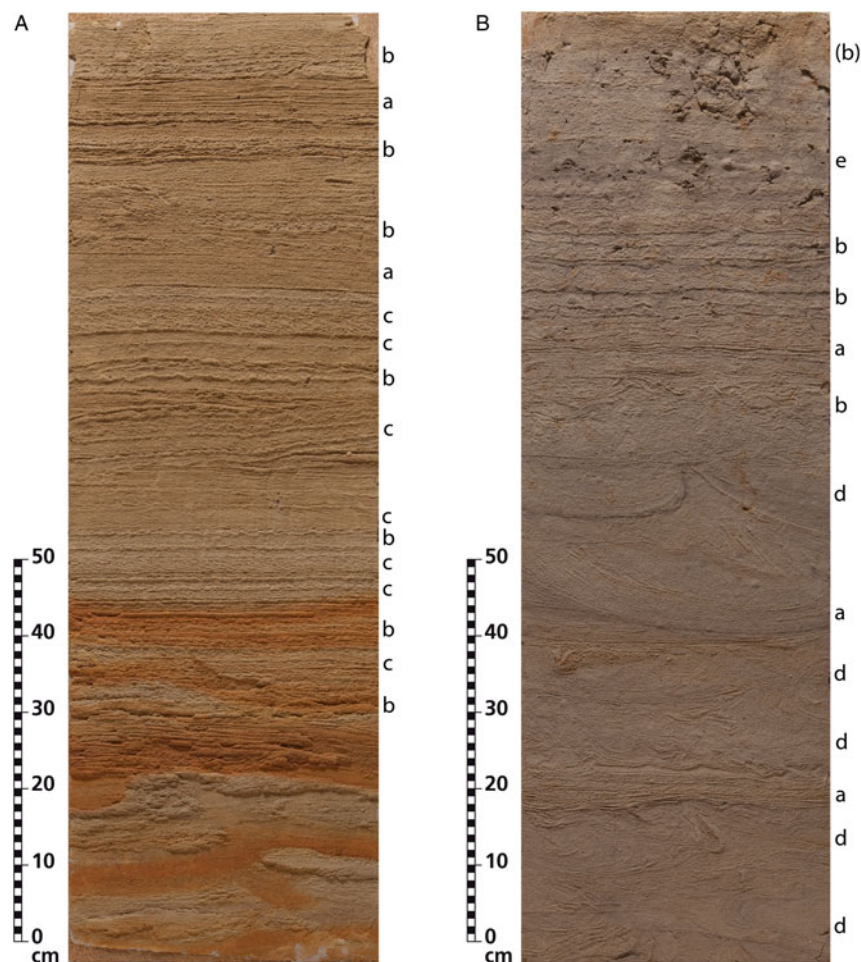


Fig. 8. (A) Lacquer peel of the base of unit 3 (Younger Coversand II) showing drying-upward sequence with wet aeolian bedding gradually changing to more dry aeolian bedding. The Usselo Soil (soil A) is situated c.0.5 m below the peel. (B) Lacquer peel of basal part of unit 5 (Drift Sand) dominated by wet aeolian bedding and involutions. Legend a: dry aeolian horizontal lamination; b: wet aeolian crinkly lamination; c: wet aeolian adhesion ripple lamination; d: involuted bedding; e: incipient soil.

et al., 2004) and 18.1 ± 2.9 ka (Fink, 2000) (Fig. 9). Based on the OSL dates and overlying radiocarbon dates from soil complex A (see below), it is concluded that unit 2 formed after the LGM during the Weichselian Late Pleniglacial and early Late Glacial. In this period, widespread aeolian deposition occurred over most of the Netherlands and surrounding regions, and the landscape was covered by an aeolian sand sheet of coversands (Older Coversand II) (Koster, 1988; Kasse, 1997, 2002; Kasse et al., 2007; Derese, 2011; Vandenberghe et al., 2013). This phase of coversand deposition lasted at least till the start of the Late Glacial, but often it continued into the Late Glacial (Van Geel et al., 1989; Kasse, 1999a, 2002). The OSL dates cannot distinguish whether the sediments belong to the Older Coversand II and/or Younger Coversand I units because the Lower Loamy Bed separating the two coversand units was not encountered. Coversand deposition in the Netherlands came to a halt during the Allerød period when the so-called Usselo Soil formed, which is a widespread soil horizon in the Late Glacial aeolian stratigraphy of northwest and central Europe (Van der Hammen 1951, 1971; Derese et al., 2009, 2012; Kaiser et al., 2009; Vandenberghe et al., 2013).

Allerød soil formation and early Younger Dryas deposition

The local pollen zones of soil complex A (Fig. 7A) can be correlated with the regionally well-established biochronostratigraphy of the Late Glacial (Hoek, 1997, 2001). The dominance of *Betula* in zone 1 (34–36 cm) followed by *Pinus* in zone 2 (29.5–34 cm) is

equivalent with respectively the *Betula* phase (PAZ 2a: 14.0–13.2 ka cal. BP) and *Pinus* phase (PAZ 2b: 13.2–13.0 ka cal. BP) of the Allerød period.

The strong decline in *Pinus* and increase of *Juniperus*, *Ericaceae* and dryland herbs in the organic, loamy fine sand (Fig. 7A: zone 3: 0–29.5 cm) is correlated with PAZ 3 (13.0–11.7 ka cal. BP) of the Younger Dryas stadial (Hoek, 1997, 2001; Bos et al., 2006). *Empetrum* and *Armeria* are typical elements of the Younger Dryas period. Pine forests declined with the climatic deterioration during the Allerød to Younger Dryas transition leading to forest fires, and a more open landscape and heliophilous vegetation developed (Van der Hammen and Van Geel 2008). Local wet conditions, as indicated by the grain size and sedimentary bedding, are reflected by high *Cyperaceae* and *Sphagnum* values. The presence of *Alnus*, *Ilex* and especially *Tsuga* probably reflects landscape instability (erosion) and reworking of pollen from the Early Pleistocene sediments below (Waalre Formation) (Kasse, 1988: pollen diagrams Korteve and Ravels). The sediments of pollen zone 3 can be regarded as the first slow response to the Younger Dryas cooling that culminated later in extensive river dune formation (unit 3).

The uncalibrated radiocarbon dates of wood fragments from soil A (Fig. 7A: pollen zone 2) ($10,870 \pm 40$; $10,900 \pm 30$ and $11,060 \pm 60$ BP; c.12.8–13.1 ka cal. BP) (Table 1) are in good agreement with the biochronostratigraphic interpretation (Allerød *Pinus* phase, PAZ 2b: 13.0–13.2 ka cal. BP; Hoek, 2001). The dates indicate a late Allerød age, and the lower part of soil complex A is

therefore correlated with the Usselo Soil (Vandenberghe et al., 2004). The Usselo Soil in the Boudewijn sand pit was previously dated at $11,240 \pm 50$ BP (c.13.1–13.3 ka cal. BP) by Schwan (1991), which is slightly older, but possibly a deeper or larger part of the soil was dated then. The dates are in accordance with the mean value of 23 dates on charcoal from the Usselo Soil in the Netherlands that cluster around 11,000 BP (c.13.1 ka cal. BP) (Hoek, 1997: 43). More recent high-resolution radiocarbon dating of charcoal from the Usselo Soil and overlying Ahrensburg occupation layer at site Geldrop-A2 gave a mean age of $10,870 \pm 15$ BP (c.12.8 ka cal. BP) (Van Hoesel et al., 2012) and $10,915 \pm 35$ BP (c.12.8 ka cal. BP) (Kasse et al., 2018) respectively.

The radiocarbon dates of the loamy, slightly organic, fine sand in the upper part of soil complex A ($10,450 \pm 190$; $11,080 \pm 180$; $10,460 \pm 180$ and $10,680 \pm 100$ BP; c.11.8–12.8 ka cal. BP) (Table 1) indicate a Younger Dryas age. This is in good agreement with the regional biochronostratigraphic interpretation of pollen zone 3 (Fig. 7A; Younger Dryas PAZ 3: 11.7–13.0 ka cal. BP; Hoek, 2001). The $11,080 \pm 180$ BP date is equal in age to the underlying Usselo Soil and is therefore considered to be too old. The gradual sedimentary transition from the fine-grained upper part of soil complex A into the overlying sand unit 3 demonstrates that deposition of unit 3 slowly started in the Younger Dryas. It shows the delayed response of the aeolian system to the Younger Dryas cooling.

Younger Dryas river dune formation (unit 3: Younger Coversand II)

Unit 3 is characterised by a coarsening-upward and drying-upward sequence with finer grained wet-aeolian deposition at the base and coarser-grained dry-aeolian deposition in the upper part (Figs 4C, D, 5 and 8A) (Schwan, 1991). The relief of the overlying podzol soil B and sporadic presence of slipface deposition points to dune formation not exceeding 5 m in height in the upper part (Fig. 4E) (cf. Schwan, 1991: Fig. 4).

The fine-grained basal part is interpreted as a wet produne environment that formed in front of the dunes that migrated over the Usselo Soil (Fig. 6B). The dune advance raised the water table in front of the dunes, fine-grained loamy material was trapped on the wet surface and organic material became preserved. In addition to system-intrinsic depositional changes, external climate factors also played a role. The cold Younger Dryas climate and vegetation decline (Fig. 7A: pollen zone 3) resulted in lower evapotranspiration and deeper seasonal frost which also increased the surface wetness. Higher lake levels have been reported for this period in eastern Noord Brabant (Bos et al., 2006).

The small-scale deformations of the Usselo Soil or peat layer (Fig. 4B) can be explained by increased frost activity at the start of the Younger Dryas period. Large involutions were found locally in the lower wet-aeolian part of unit 3 (Fig. 4F). Metre-scale and widespread loading phenomena in Weichselian Late Pleniglacial deposits are interpreted as the result of melting of ice-rich permafrost (Vandenberghe, 2006). Local permafrost conditions during the Younger Dryas have been reported in the Netherlands (Kasse, 1995; Isarin, 1997). However, neither ice-wedge casts nor frost cracks were found in unit 3. Therefore, these local large-scale involutions are not interpreted as periglacial loadings, but they probably resulted from deposition of sand on a water-saturated surface leading to compaction, oversaturation, quicksand formation and involution (Schwan, 1990; Kasse, 1999b).

The radiocarbon dates from the upper part of soil complex A (c.11.8–12.8 ka cal. BP) are in good agreement with OSL ages of 12.2 ± 0.7 , 13.1 ± 0.9 and 11.5 ± 0.7 ka from the base of unit 3 obtained by Vandenberghe et al. (2004) and with OSL ages between 10.0 ± 1.5 and 15.1 ± 3.9 ka from unit 3 by Fink (2000) (Fig. 9). All dates combined indicate that unit 3 was mainly deposited in the (second part of the) Younger Dryas period between c.12.3 and 11.7 ka. These results are in accordance with previous research in which the formation of the Younger Coversand II unit was placed in the (second half of the) Younger Dryas period (Bohncke et al., 1993; Kasse et al., 1995, 2018; Isarin et al., 1997; Derese et al., 2010b, 2012).

The Younger Coversand II at Ossendrecht is part of an extensive, 4–8 km wide, dune belt (Fig. 1: drift sand and river dunes). The south–north orientation of this dune belt is parallel to the former Late Glacial Scheldt valley that is present in the subsurface below the young Holocene polders (Kiden, 1995; Westerhoff & Dobma, 1995; Vos and Van Heeringen 1997). Sand was blown out of the Scheldt valley during the Younger Dryas stadial and therefore these dunes are interpreted as source-bordering river dunes on the right bank of the Scheldt. The relatively coarse grain size and high amount of grains larger than 500 μm (Fig. 5) indicate limited sorting due to a short transport distance of the fluvial source material. The parabolic dune forms indicate a west-southwesterly wind direction and trapping by vegetation. Despite the Younger Dryas cooling, vegetation cover did not disappear, although its composition changed (Fig. 7A). This is in accordance with reconstructions and climate modelling experiments for northwestern Europe (Isarin et al., 1997). Similar dunes have been described on the east bank of the Maas valley (Bohncke et al., 1993; Kasse, 1995; Kasse et al., 1995; Hoek et al., 2017; Woolderink et al., 2019) and in the Rhine–Meuse fluvial area of the central and western Netherlands (Pons, 1957; Hijma et al., 2009). Increased aeolian activity, particularly during the second phase of the Younger Dryas, has been ascribed to a delayed response to cooling, increased aridity and vegetation decline (Hoek, 1997; Isarin & Bohncke, 1999; Kasse, 1999a, 2002). A decline of the forest vegetation cover (Fig. 7A) and a change from meandering to braided rivers are held responsible for this new phase of aeolian deposition, which is therefore more regional than the supraregional extensive Older Coversand II / Younger Coversand I. East of the river dune belt, Younger Coversand II deposition did not occur and the Late Glacial Usselo Soil was probably incorporated later in the Holocene soil (Kasse, 2002; Kasse et al., 2018).

Early and Middle Holocene soil formation (podzol soil B)

River dune formation (unit 3, Younger Coversand II) does not seem to continue over the Younger Dryas to Holocene boundary (11.7 ka cal. BP), given the uncertainty of the OSL dates (Fig. 9). This suggests no or a limited time lag between the climate warming and aeolian activity. Aeolian transport stopped and the dunes were stabilised by rapid colonisation of the surface by nearby vegetation (herbs and birch) at the start of the Holocene (Hoek, 1997, 2001). Soil formation started during the early Holocene, and over time the humic-iron podzol soil B developed on the well-drained dunes and peat was formed in former dune depressions (Schwan, 1991).

The dominance of *Corylus* and *Pinus* in the lower part of the podzol soil B (Fig. 7B: zone 1: 23–25 cm) points to formation during the Boreal period (9000–8000 BP; 10.8–8.9 ka cal. BP) (Van Geel et al., 1981). The high Ericaceae values indicate the early presence of heathlands, which is in accordance with recent findings for

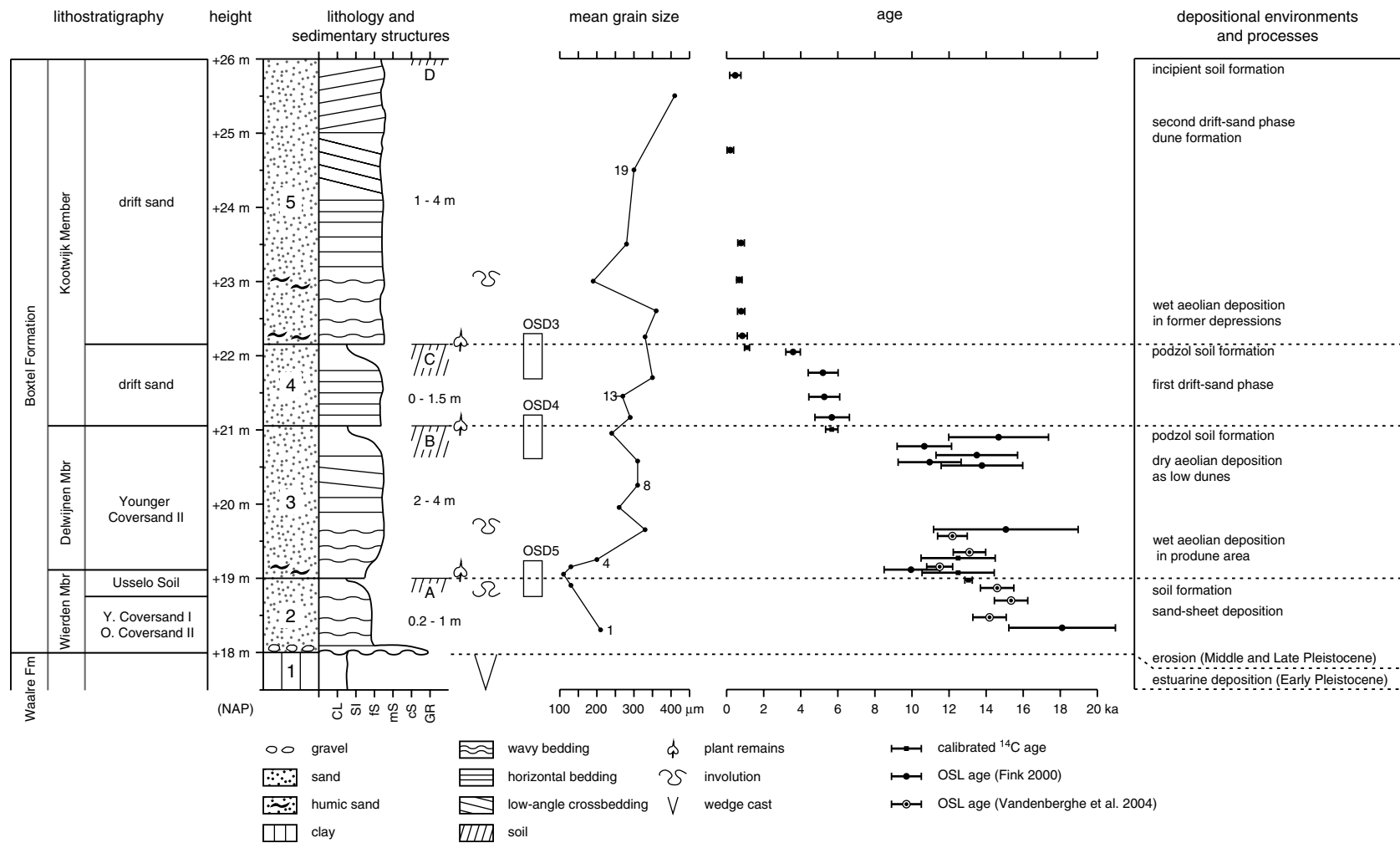


Fig. 9. Synthesis of the Late Glacial and Holocene lithostratigraphy, aeolian depositional phases, soil formation and timing of the environmental changes. 1 to 5 are lithological units; A–B–C–D are soils; OSD3–4–5 are pollen profiles (Fig. 7); 1, 4, 8, 13, 19 are grain-size samples (Fig. 5B). Note that there is almost no time gap between the radiocarbon-dated peaty top of the soils and the OSL-dated overlying sands. Age jumps in the time–depth plot coincide with soil formation (sedimentary hiatus), and high sedimentation rates are present in the aeolian units. Calibrated radiocarbon ages of the Usselo Soil (soil A) are in good agreement with OSL dates of units 2 and 3 from different publications.

the central Netherlands (Sevink et al., 2013, 2018). The continuous presence of *Alnus*, *Quercus*, *Tilia*, *Ulmus* in zone 2 (Fig. 7B; zone 2: 7–23 cm) is characteristic for the Quercetum Mixtum of the Atlantic period (8000–5000 BP; 8.9–5.8 ka cal. BP). The presence of *Fraxinus* indicates the second half of the Atlantic period (Van Geel et al., 1981).

The radiocarbon dates of the peaty top of podzol soil B (4940 ± 50 and 4940 ± 60 BP (c.5.6–5.8 ka cal. BP; Table 1) indicate a late Atlantic to early Subboreal period which is in agreement with the biostratigraphic interpretation. The OSL dates of the underlying unit 3 and the radiocarbon dates of the top of podzol soil B show that soil B formed during a long period of landscape stability in the Early and Middle Holocene, between c.11.7 and 5.8 ka cal. BP (Fig. 9). Podzol soil B grades locally into a peat layer in former river dune depressions. The peat was dated between 9050 ± 45 and 3000 ± 30 BP (c.10.3–3.1 ka cal. BP) by Schwan (1991), which indicates that the peat and podzol soil B formed simultaneously. Podzol soils of similar age were described by Sevink et al. (2013: soils S1, S2, S3 combined) and Van Mourik et al. (2010). It shows that the onset of podzolisation and development of ericaceous vegetation occurred prior to the introduction of Neolithic farming, which is earlier than previously assumed (Louwe Kooijmans, 1995: 419) and in accordance with recent investigations (Sevink et al., 2013, 2018; Doorenbosch and Van Mourik 2016; Sevink and Van Geel 2017). Mineralogical differences in parent material may be responsible for differences in the timing of the onset of acidification and podzolisation (Kluiving et al., 2015). Aeolian quartz-rich dune sands, as in the investigated site, were more susceptible to rapid acidification and podzolisation than loamy coversands and fluvial deposits in which the acidification took place at a later stage.

Middle Holocene drift-sand deposition (unit 4)

The areal distribution of unit 4 is limited over a short distance of c.100 m (Fig. 6A, D). Grain size and sedimentary bedding resemble unit 3 (Younger Coversand II) (Figs 5 and 9). However, unit 4 is interpreted as a local drift-sand deposit, as it is intercalated between two Holocene podzol soils B and C (Fig. 6D). The distinction between coversands and drift sands is based on climate and chronostratigraphy and refers to cold-climate Pleistocene and Holocene aeolian sands respectively (Koster, 1982). Unit 4 was probably formed by local reworking of the underlying river dunes. The local occurrence of unit 4 was previously also established by Schwan (1991, his Fig. 4). He recorded 450 m of quarry pit faces, and generally one Holocene podzol soil was observed. Only in a short 30 m section (165–194 m) were two podzols found.

Drift sand unit 4 overlies podzol soil B (4940 ± 50 and 4940 ± 60 BP; c.5.6–5.8 ka cal. BP; Table 1), possibly indicating Middle Holocene drift-sand deposition. However, indirect dating of sand units by radiocarbon dating of underlying soils may give unsatisfactory results (Van Mourik et al., 1995, 2010, 2011; Koster, 2009; Wallinga et al., 2013; Pierik et al., 2018). The organic material of soil B has accumulated over the Early and Middle Holocene periods, and resistant carbon fractions of the soil-forming period are still incorporated in the soil. Therefore, radiocarbon dating of buried soils gives a maximum age of the overlying sand deposit and does not necessarily represent the start of sand deposition (Vandenbergh et al., 2013; Pierik et al., 2018). To circumvent this problem the gradual peat-to-sand transition at the top of soil B, that reflects sand being trapped in the sedge vegetation, was sampled for radiocarbon dating.

The radiocarbon dates of the top of soil B are 4940 ± 50 and 4940 ± 60 BP (c.5.6–5.8 ka cal. BP). OSL dating of overlying unit 4 revealed a mean age of 5449 ± 797 ($n = 3$) (Fink, 2000) (Fig. 9). The radiocarbon dates of the top of soil B and the OSL dates of unit 4 (c.5500 years) are therefore in good agreement. This indicates that radiocarbon dates from the top of peaty soils can be used to reliably date the onset of overlying sand deposition (unit 4). The three OSL dates of unit 4 have overlapping ages, which indicates a short phase of drift-sand deposition at c.5500 cal. BP at the Atlantic to Subboreal transition during the Middle Neolithic (Fig. 9). A similar drift-sand phase was reported by Zonneveld (1965) from the nearby Kalmthoutse Heide, c.4 km east of the Boudewijn pit, which was dated later than 5300 ± 90 BP. The cause for the first drift-sand phase (unit 4) at c.5500 cal. BP is not clear. The vegetation reconstruction of the underlying podzol soil B shows the presence of heathlands (Ericaceae) and an increase of grasses and sedges indicating a local decrease of the forest cover and opening of the vegetation, possibly by local forest clearing (Fig. 7B). Dryland herbs like *Artemisia*, Chenopodiaceae, Asteraceae and Brassicaceae can indicate agricultural activity (Bos & Zuidhoff, 2015: 51). However, Cerealia pollen, indicative for arable land, has not been found. The pollen assemblage of soil B and the first drift-sand deposition might reflect the impact of early agriculture on the vegetation in the region during the Neolithic. However, it was stated that the impact of farming on the vegetation and on the landscape was still limited or local during the Late Neolithic (Louwe Kooijmans, 1995: 419; Bos & Zuidhoff, 2015: 49; Broothaerts et al., 2015; Van Beek et al., 2015) while it increased during the Bronze Age. Neolithic occupation and a first increase in Cerealia pollen is reported in the Maas valley in the southeastern Netherlands from c.6400 cal. BP (4365 BC; Bos & Zuidhoff, 2015). According to Sevink et al. (2013: 260) the major drift-sand phase D-3 around 5000 cal. BP in the central Netherlands coincides with the first intensive land use in this region, suggesting a causal relation between this early agricultural phase, land degradation and sand-drifting.

Middle Holocene drift-sand phases have been reported from the Netherlands and Germany (Van Mourik et al., 1995, 2010, 2012; Tolkendorf & Kaiser, 2012; Sevink et al., 2013, 2018; Koster, 2017; Pierik et al., 2018). The causes for these earlier Meso- and Neolithic drift-sand phases, whether by climate change (cooling, aridity, storminess), human activity (burning of vegetation, agriculture, overgrazing) or natural soil degradation, is still debated (Jungerius & Riksen, 2010; Tolkendorf & Kaiser, 2012; Sevink et al., 2013, 2018; Kluiving et al., 2015; Pierik et al., 2018).

Holocene climatic oscillations were probably not strong enough to cause landscape degradation and to initiate sand-drift formation. Although the vegetation composition changed (Hoek & Bos, 2007), the protective vegetation cover persisted, hampering the initiation of deflation and drift-sand formation. On the other hand, the impact of man on the landscape increased from the Neolithic period onwards. Our results confirm that deforestation and soil cultivation may have led to local barren surfaces prone to local sand drifting in the Neolithic.

The occurrence of drift-sand phases need not be time-equivalent in the Netherlands. Temporally and spatially different responses may have occurred to the same forcing depending on the carrying capacity and resilience of the landscape. Differences in parent material such as mineralogy and loam content resulted in differences in vegetation and soil formation (Kluiving et al., 2015). On quartz-rich parent materials with poorly buffered soils, the vegetation had a lower regenerative capacity following human impact, and

the surface was more susceptible to deflation and drift-sand formation than on the richer soils. First small-scale sand drifts seem to have occurred on nutrient-poor, depleted coversands and river dunes (Willemse & Groenewoudt, 2012; Sevink et al., 2013, 2018) similar to the investigated site Boudewijn in western Noord Brabant. The same forcing (e.g. small-scale slash-and-burn agriculture) may therefore have resulted in a site-specific environmental response and a different threshold towards deflation.

Middle and Late Holocene soil formation (podzol soil C)

Podzol soil C developed in the upper part of unit 4 and is therefore younger than c.5500 years. The presence of *Jasione* at the base of pollen zone 1 indicates dry open sandy soil, following the deposition of unit 4 (Fig. 7C). The presence of *Alnus*, *Corylus*, *Quercus*, *Tilia* and *Ulmus* and absence of *Fagus* and *Carpinus* in zone 1 indicate a late Atlantic or early Subboreal age (c.5.8–4.0 ka) (Janssen, 1974; Bos & Zuidhoff, 2015). The decline of *Ulmus* in the top of zone 1 possibly indicates the start of the Subboreal. The presence of high Ericaceae values and dryland herbs like Chenopodiaceae, Asteraceae and Brassicaceae may indicate agricultural activity; however, *Cerealia* pollen has not been found in zone 1. Pollen zone 1 of soil C (Fig. 7C) resembles zone 2 of the underlying podzol soil B (Fig. 7B). Since unit 4 has been deposited in a short period around 5500 years ago, it is very likely that the two pollen zones are closely connected in time. This means that little time is present in clastic unit 4 (high sedimentation rate), while most of the time is represented by soil formation (soils B and C).

The presence of *Fagus* and *Carpinus* in zone 2 of soil C (Fig. 7C) indicates a late Subatlantic age (younger than c.2000 cal. BP) (Janssen, 1974; Bos & Zuidhoff, 2015). The low values of *Alnus*, *Betula*, *Corylus*, *Quercus* and *Ulmus* in combination with high Ericaceae values point to deforestation and heathland development. The presence of *Cerealia* type is an indication for agriculture, and weeds related to agriculture (Chenopodiaceae, Asteraceae, Brassicaceae, *Rumex* and *Plantago*) are present in the pollen diagram. The occurrence of *Centaurea* type is generally associated with the Middle Ages and is found from the 10th to 11th century AD onwards (Bakels, 2012; Bos & Zuidhoff, 2015: 30; Sevink and Van Geel 2017: 285).

The radiocarbon dates from the peaty top of podzol soil C of 1000 ± 30 and 890 ± 30 BP (c.1000–800 cal. BP) (Table 1) indicate that the final stage of soil formation occurred during the Middle Ages or later. This age, in close agreement with the pollen analytical results presented above, shows that radiocarbon dates from the peaty top of a soil can be reliable.

Based on the OSL dates of the underlying drift sand unit 4 (c.5500 years) in combination with the radiocarbon dates and vegetation reconstruction of soil C, it is concluded that podzol soil formation (soil C) covered a 4500 year period from c.5500 to 1000 cal. BP (Fig. 9).

Late Holocene drift-sand deposition (unit 5)

Unit 5 is a widespread drift-sand unit observed in all pit faces (Fig. 3). The grain-size characteristics resemble units 3 and 4, and the coarsest intervals are observed in the top of the unit (Figs 5 and 9). The coarsening-upward trend in combination with a change from wet to dry aeolian bedding types reflects a drying-upward sequence similar to unit 3 (Fig. 8) (Schwan, 1991: 161). Initial planation of older dunes, erosion of the podzol soil and filling of dune depressions under wet conditions was followed by dry-

aeolian dune formation (Figs 3C, 6C). The finer grain size and faint humic laminations at the wet-to-dry transition may represent a short stand-still phase in the sedimentation and associated initial soil formation (cf. Schwan, 1991: hydromorphic soils in Kootwijk Formation).

Decimetre-scale involutions especially at the transition from the wet to dry aeolian deposits are attributed to the deposition of dry sand on a water-saturated surface leading to a reversed density gradient, oversaturation and load casting (Figs 4G and 8B). Successive involuted levels indicate successive phases of deposition, groundwater table rise and load casting (Schwan, 1991: 165).

Unit 5 overlies podzol soil C. The pollen assemblage of the upper part of soil C (Fig. 7C) demonstrates that unit 5 was formed during or following the Middle Ages (Bos & Zuidhoff, 2015: 30). The presence of the weakly developed present-day soil indicates that deposition continued until recently. Many active drift sand areas in western Noord Brabant were reforested from the end of the 19th century onwards (Zonneveld, 1965). The radiocarbon dates from the top of the underlying podzol soil C (1000–800 cal. BP) are in good agreement with the OSL dates of unit 5 ranging between 0.9 and 0.3 ka (Fink, 2000) (Fig. 9). The dates indicate no or limited time lag between the end of soil formation and the start of aeolian deposition. Radiocarbon dating of soils can give good results provided that the top of wet-type peaty soils is sampled. A reliable chronology of aeolian deposition can be established based on both radiocarbon dating and pollen analysis of soils in combination with luminescence dating of sand units.

Medieval to recent drift-sand phases have been reported previously from the nearby Kalmthoutse Heide, c.4 km east of the Boudewijn pit (Zonneveld, 1965: drift sand later than 1310 ± 60 BP), from eastern Noord Brabant (Van Mourik et al., 2010; Kasse et al., 2018) and from adjacent Belgium (Beerten et al., 2014). Extensive drift-sand formation in the Netherlands and northwestern Europe occurred especially from the late Middle Ages onwards (c.1100 cal. BP) (Castel et al., 1989; Koster et al., 1993; Koster, 2009, 2010; Vandenberghhe et al., 2013; Pierik et al., 2018: Fig. 11).

This young drift-sand phase is generally ascribed to human occupation. Deforestation and expansion of agriculture, grazing and burning practices, formation of roads and cattle/sheep drifts, and the use of plaggen as fertiliser led to the remobilisation of the Late Weichselian aeolian sands (Koster, 2009: 100; Pierik et al., 2018). The palynological indications for deforestation and agriculture in the upper part of podzol C support this view (Fig. 7C). A climatic cause for drift-sand formation, as has been suggested previously (Heidinga, 1984, 2010; Jungerius & Riksen, 2010), cannot be confirmed, although drift-sand expansion may have intensified during the Little Ice Age as stated by Pierik et al. (2018).

Conclusions

A stacked sequence of Late Weichselian and Holocene aeolian sediments and soils in the southern Netherlands reveals autogenic changes in the sedimentary sequence and the allogenic impact of climate and man. Coarsening and drying-upward trends are caused by system-intrinsic changes in topography and drainage conditions related to the accumulation of wind-blown sand and dune migration. Aeolian deposition changed over time from supraregional (Older Coversand II and Younger Coversand I) to regional (river dunes / Younger Coversand II) to local (Drift Sand phases).

Climate-controlled sand accumulation (Older Coversand II and Younger Coversand I) occurred during the Late Pleniglacial and early Late Glacial. A soil complex formed during the Allerød (Usselo Soil) and early Younger Dryas. Vegetation cover strongly changed (decline of pine, increase of herbs) at the Allerød – Younger Dryas transition, but the aeolian response was limited. Younger Coversand II deposition occurred during the second part of the Younger Dryas stadial (c.12.3–11.7 ka cal. BP), implying a delayed response to the Younger Dryas cooling. Source-bordering parabolic river dunes formed along the Scheldt valley by southwesterly winds. Aeolian deposition stopped around the Younger Dryas to Holocene transition, implying a limited time lag to the climate change and rapid vegetation response.

Early podzolisation and heathland formation occurred in the Early and Middle Holocene, between c.11.7 and 5.8 ka cal. BP, prior to the introduction of Neolithic farming, as a consequence of natural or anthropogenic soil degradation. In the Middle and Late Holocene a second podzol soil formed between c.5500 and 1000 cal. BP. In line with earlier studies in the Netherlands and adjacent Belgium, we find a small-scale drift-sand phase around 5500 years ago, possibly related to agricultural activity and a more extensive (post-)Medieval drift-sand phase after 1000 cal. BP caused by deforestation and agriculture.

Climate was the dominant forcing factor during the Late Weichselian, while forcing by man increased and became dominant during the Holocene. The study shows a good correspondence and limited hiatuses between the radiocarbon-dated top of peaty soils and luminescence ages of overlying sand units. A multi-proxy approach based on sediment characteristics, pollen analysis and independent dating control enables robust landscape reconstructions.

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